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RAINDROP SIZE-DISTRIBUTION IN HAWAIIAN RAINS

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ABSTRACT

A brief survey of the major techniques of raindrop size-sampling is given. The filter-paper technique, finally adopted for use in this study, adapts itself admirably to the sampling of Hawaiian orographic rains. The change in the drop-size distribution of rain as it falls from cloud to ground may be considerable. It is affected by wind shear, gravity separation, evaporation and drop collision. The evaporation error alone can be appreciable. The many small drops of the Hawaiian orographic rains may completely evaporate in a sub-cloud fall of only 1000 m. The evaporation problem was eliminated, and the others minimized, by sampling all the orographic rain at cloud base or within the cloud itself.

Drop-size distributions were obtained in such non-orographic rains as thunderstorms and cyclonic storms. The pertinent meteorological parameters, such as liquid-water content, median drop diameter, and radar reflectivity, agree reasonably well with the values given by other investigators.

The measurements made in orographic rains from non-freezing clouds, however, lead to considerably different values of these factors. The raindrop distributions are narrow, with the largest drops rarely exceeding 2 mm in diameter. In general, the higher the intensity, the more numerous are the drops at the large end of the spectrum. At the small end of the drop spectrum (<0.4 mm), however, increased intensity is accompanied by a decrease in the drop count. Distributions of this type indicate the absence of any chain-reaction process.

Concentrations of drops less than 0.5 mm in diameter often are in excess of 40,000 m^{-3}. These large numbers of small drops give low values for median drop diameter and radar reflectivity, but high values of liquid-water content.

All of the drop distributions have been put into three categories: (1) non-orographic rain, (2) orographic rain at cloud base, and (3) orographic rain within the cloud and near cloud top. In each case, regression equations have been developed to express the meteorological parameters as a function of rain intensity.

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1. Introduction

In October 1951, the writer and Mr. A. H. Woodcock went to the Hawaiian Islands to begin a joint ten-months study with the Meteorology Department, Pineapple Research Institute and Hawaiian Sugar Planters' Association. The study was aimed toward a better understanding of the basic mechanism of warm-cloud rain. It is believed that large salt particles of marine origin form the nuclei from which raindrops develop, first by condensation and later by accretion (Woodcock, 1952). To test this hypothesis further, three separate programs of study were carried out: (1) measurements were made of the air-borne salt particle distribution beneath, at and above the cloud layer, (2) the variation of rainwater chloride-content *vs.* intensity was studied, and (3) the raindrop size-distributions at various points within the cloud were obtained.

Inasmuch as the meteorological literature, with the exception of Anderson's (1948) work, contains little information on raindrop size-distributions from non-freezing clouds, it was felt that the data obtained in connection with this third program would be of of sufficient interest to warrant publication. It is becoming increasingly more evident that a study of the drop distribution may enable us better to understand the growth mechanism of the drops.

One of the earliest papers on raindrop size described observations of splash pattern on slates (Lowe, 1892). At about this time, the idea of exposing chemically treated filter papers to the rain was suggested, but it remained for Wiesner (1895) to publish the first detailed results. A novel approach to raindrop size-measurements was achieved with the flour technique (Bentley, 1904). The raindrops, on falling into a flour-filled container, produced hard dough pellets whose size was a function of the diameter of the original raindrops. This method has subsequently been used by several investigators (Laws and Parsons, 1943; Chapman, 1948; Blanchard, 1949a). An account of European investigations of raindrop size and accompanying instrumentation prior to 1942 can be found in an excellent survey paper by Neuberger (1942).

In an effort to develop a drop-size measuring technique which would eliminate the splashing and spreading of the large drops on contact with the sampling surface, the writer (Blanchard, 1949b) experimented with soot-coated 100- and 50-mesh brass screens. Raindrops, in passing through the screen, removed a circular area of soot whose diameter

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was a function of the drop size. This method was considerably improved when nylon screens were substituted for wire screens (Mt. Washington Observatory, 1951a). The nylon screens were treated with a petroleum ether-lanolin solution and then covered with powdered sugar. In this manner, some excellent raindrop samples have been obtained. Mr. Woodcock recently attempted to use these screens from aircraft flying at speeds of 60–80 mi/hr. With low speeds and low relative humidities, a drop-size distribution can be obtained; but in the high humidity region near cloud base, and within the rain area, the hygroscopic sugar particles absorb water and render the screen useless. It would appear, from some brief experiments in sooting nylon screens, that the hydrophillic soot particles from acetylene smoke would serve in lieu of powdered sugar for measurements of drop size from aircraft.

Electronic techniques have been developed in an attempt to obtain continuous measurements of drop size in flight. Cooper (1951) has used a balloon-borne instrument for telemetering raindrop size. An instrument, similar in principle, has been used in France (Maulard, 1951). In the United States, a number of reports, dealing with both optical and momentum devices, have been issued on air-borne instrumentation (Katz, 1952). At the time of this writing, few of these instruments have been put into use.

In Australia, a raindrop spectrograph has been used to obtain continuous drop-size measurements at the ground (Bowen and Davidson, 1951). This ingenious and relatively simple technique permits a direct determination of raindrop size.

2. Hawaiian climate

As any study of this type should be made with cognizance of the influence of the local topographical and meteorological conditions, a brief discussion of these factors and their influence on Hawaiian rainfall will be given.

The eight Hawaiian Islands, some 2400 mi southwest of San Francisco, are oriented northwest-southeast and extend from a latitude of 19 to 22°N. The entire island chain is located within the Pacific northeast trades. These trades are characterized by a temperature inversion with a modal elevation of 6000 ft. Below the inversion, the air is moist and turbulent with an average lapse rate of 8.3C/1000m. As one passes up through the inversion, the air becomes quite dry and free from turbulence. The usual convective and orographic clouds are normally limited by the inversion. It is only on the relatively infrequent occasions when the trade winds are weak or subside completely that the clouds remain over the islands for a sufficient time to build up convectively to high altitudes. As these conditions are so infrequent, it has proved difficult to evaluate properly the results of dry-ice seeding in Hawaii (Leopold and Mordy, 1951).

A marked departure from the normal trade-wind weather is introduced by the passage of easterly waves in the trade-wind current and by the Kona storm (Simpson, 1952). The Kona storms, occurring perhaps 2–3 times during the winter and spring, are cyclonic storms which develop to the northwest of Hawaii. During the day or two of Kona-type weather, heavy rainfall is experienced throughout the islands.

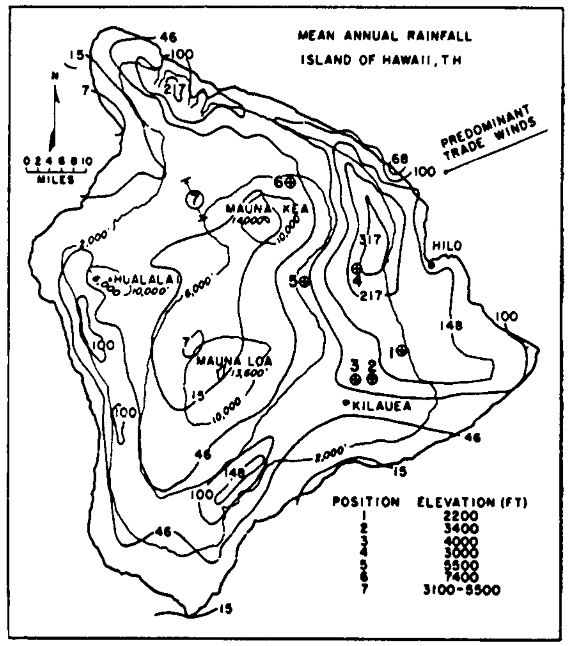

FIG. 1. Isohytal map of island of Hawaii, with location of the seven sampling positions.

The topography of the islands is the major factor in the formation of the orographic clouds. This is effectively shown in the isohyets of the annual rainfall, especially those of the island of Hawaii (see fig. 1). Strong isohyetal gradients are set up in critical areas of trade-wind flow. For example, note the marked increase in annual rainfall from sea level to a point some 10 mi up the east flank of Mauna Kea. In this distance, the annual rainfall increases by 250 in. A rapid decrease of annual rainfall with altitude is found at higher elevations. An explanation for this rainfall maximum has been given by Leopold (1949), who attributes it to the splitting of the trade winds by the huge volcanic cones. He states: "Streamlines drawn in accordance with the observed splitting of the trades by each of the two cones, Mauna Loa and Mauna Kea, would converge directly over the observed zone of greatest rainfall."

TABLE 1. Summary of 113 rain samples made during study.

Sample no.	Date	Time (Hawaiian Standard)	Position	R (mm/hr)	W (mg/m³)	Z (mm⁶/m³)	d_0 (mm)	Number of drops per cubic meter within 0.2-mm size interval centered about indicated size (mm)																			
								0.1	0.3	0.5	0.7	0.9	1.1	1.3	1.5	1.7	1.9	2.1	2.3	2.5	2.7	2.9	3.1	3.3	3.5	3.7	3.9
1	3-27-52	0814	6	1.6	230	76	0.39	11,200	5,800	590	172	80															
2	3-27-52	0842	6	3.5	181	2,112	1.3	*	250	100	44	24.5	43	15	13	9.2	5.7	—	2.6	—	2.35						
3	3-27-52	0845	6	3.1	148	2,285	1.75	*	182	25	31	25	14	6.1	7.3	6.7	1.6	7.4	2.85	1.37	1.3						
4	3-27-52	0858	6	2.95	154	1,668	1.69	*	248	92	49	25	16	8	7.4	11.8	7.8	3	1.4	1.4							
5	3-27-52	0908	6	4.4	220	2,980	1.75	*	315	125	42	32	22	9.7	22	2	11.3	3.55	5	1.62	1.52						
6	3-27-52	0914	6	5.5	324	1,954	1.28	*	630	260	158	109	60	35	41	4.8	8.8	—	4								
7	3-27-52	1118	6	2.9	165	1,130	1.41	*	191	110	35	32.5	37.5	20.8	15.5	8.6	7.9	—	1.2								
8	3-27-52	1120	6	2.8	166	1,039	1.36	*	430	23.5	36	53	41	17	20	4.7	5.8	2.7									
9	3-27-52	?	6	2.7	174	758	1.17	*	420	162	45	82	43	30.5	13	3	5.5										
10	3-27-52	1412	6	1.7	183	132	0.55	5,700	1,500	900	162	61	36.5														
11	3-27-52	1420	6	1.4	147	279	0.47	8,100	2,600	430	59	26	16.8	9.6	—	—	3.6										
12	3-27-52	1640	6	2.5	154	960	1.3	*	520	252	72	47	18.3	3.2	14.5	7.9	2.5	4.7									
13	3-27-52	2025	6	0.28	31	21	0.55	521	471	100	28.5	10.2	5.75														
14	1-19-52	1031	8	2.9	185	1,241	1.19	9,300	700	120	50	56	50	22	15	8	—	—	1.7	1.6							
15	1-19-52	1056	8	1.8	95	811	1.42	2,200	160	39	9.2	17	15.5	16	8.6	5.7	2.1	—	0.94	0.9							
16	1-19-52	1108	8	7.0	325	8,210	1.88	14,000	560	130	49	7.6	46	5.8	15.5	14.5	8.8	4.2	—	3.8	3.7	—	—	3.4			
17	1-19-52	1135	8	10.7	468	13,040	2.26	11,000	1,050	82	34	55	19	—	29	13.5	—	8.8	11.5	2.7	10	—	2.45	2.4			
18	1-19-52	1150	8	127	5,330	153,000	2.32	*	*	1,650	320	165	185	124	135	34	73	170	230	63	61	33	8.2	24			
19	1-19-52	1514	8	6.7	334	4,090	1.68	1,240	230	94	31	37	52	46	12	3.7	21	10	6.2	3							
20	1-19-52	1516	8	12.1	524	12,790	2.04	1,600	225	41	48	11	23	12	44	20	22	12	11	2.7	7.7	—	—	—	2.3		
21	1-19-52	1525	8	20.8	851	18,480	2.0	460	190	94	77	54	70	15	36	38	59	15	28	6.7	3.2	—	—	—	3.1		
22	1-19-52	1533	8	17.5	764	12,990	1.88	1,500	700	150	39	68	31	50	70	42	24	17	22	9	—	—	2.9				
23	2-11-52	1354	8	0.8	33	434		—	—	—	—	1.5	0.7	0.5	3.1	1.4	4	1.3	0.4								
24	2-11-52	1355	8	3.6	146	6,000		—	—	2.8	—	1.8	1.5	1.4	—	1.15	1.02	1	2.8	2.7	4.4	1.5	1.44	0.7			
25	2-11-52	1356	8	1.25	57	885		—	—	2.1	10.5	4.1	2.3	1.0	2.8	2.5	3.2	3	1.4	0.7							
26	2-11-52	1401	8	8.8	470	16,300	2.0	2,900	360	120	54	24	10	14	32	22	17	3.3	3.1	6	8.6	—	—	—	—	—	2.5
27	2-11-52	1402	8	27.5	1,170	23,440	2.05	11,200	680	230	55	120	66	65	44	34	50	36	17	27	10	5					
28	2-11-52	1404	8	12.0	640	7,680	1.65	2,400	480	86	13	25	21	52	70	50	32	5.4	2.6	4.8	4.6						
29	2-11-52	1409	8	5.25	293	3,630	1.42	3,600	690	210	95	105	26	26	20	15.5	8.5	—	2.6	—	—	—	2.2				
30	2-11-52	1419	8	2.9	154	2,330	1.71	3,500	140	13.5	5.6	29	—	1.6	19	4.1	5	3.6	1.15	1.1	1.04	1					
31	7-8-52	1423	1	1.5	156	73	0.59	—	440	970	300	43															
32	7-8-52	1433	1	3.4	351	189	0.59	830	2,240	1,880	530	139	13														
33	7-8-52	1437	1	9.2	610	2,220	1.05	—	1,640	356	510	310	160	93	10.3	28.8	8.8										
34	7-8-52	1442	1	3.9	379	235	0.64	—	850	2,000	660	208	9.3														
35	7-8-52	1445	1	6.2	550	673	0.74	730	2,000	2,050	720	153	180	31													
36	7-8-52	1451	1	7.3	522	1,114	0.98	—	1,010	472	236	464	240	35	20.4												
37	7-8-52	1509	1	2.6	235	273	0.62	—	592	1,320	282	79	17	15	6.6												
38	7-8-52	1517	1	6.5	494	870	0.89	—	340	920	660	208	235	32	7.3												
39	7-8-52	1526	1	2.8	338	314	0.69	—	1,150	1,280	535	175	56	8.1													
40	7-8-52	1539	1	3.8	398	173	0.52	—	1,740	3,100	450	92	13.2														
41	7-8-52	1545	1	1.8	196	58	0.46	122	2,100	1,670	180	17.7															
42	7-8-52	1605	1	5.1	448	473	0.76	—	870	1,460	620	375	71	10													
43	7-8-52	1617	1	8.7	682	1,080	0.92	—	740	1,240	680	490	320	31													
44	7-8-52	1629	1	5.3	652	230	0.47	245	7,400	6,000	300	18.2															
45	7-8-52	1655	1	0.31	79	5	0.22	20,000	2,700	28	23																
46	7-8-52	1714	1	5.05	660	200	0.47	75,000	6,600	4,200	460	146															
47	7-8-52	2142	4	13.3	1,243	1,400	0.73	45,000	10,900	2,200	1,700	540	332	55													
48	7-8-52	2146	4	3	330	140	0.57	1,600	3,400	1,650	520	98															
49	7-8-52	2150	4	3.6	369	257	0.62	4,700	4,600	810	635	177	23	6.7													
50	7-8-52	2155	4	4.7	672	154	0.41	29,200	13,700	4,000	248	98															
51	7-8-52	2203	4	1.8	298	47	0.39	9,000	9,400	1,030	152	10.8															
52	7-8-52	2208	4	8.2	816	531	0.61	4,600	2,480	4,400	1,400	280	83														

* Splashing of large drops by wind prevented accurate determination of the concentration of these drops. It is believed that the count would have been <500 m⁻³.

TABLE 1. (Continued)

Sample no.	Date	Time (Hawaiian Standard)	Position	R (mm/hr)	W (mg/m^3)	Z (mm^6/m^3)	d_0 (mm)	Number of drops per cubic meter within 0.2-mm size interval centered about indicated size (mm) 0.1	0.3	0.5	0.7	0.9	1.1	1.3	1.5	1.7	1.9	2.1
53	5- 6-52	1648	4	1.02	160	19.6	0.4	2,270	4,200	1,058								
54	5- 6-52	1654	4	1.5	196	60	0.5	3,870	3,210	890	287	19.8						
55	5- 6-52	1710	4	0.95	149	20	0.39	4,800	4,050	790	41							
56	5- 6-52	1718	4	0.51	76	10.2	0.41	2,000	1,600	580								
57	5- 6-52	1721	4	2.6	262	168	0.71	2,000	2,170	524	590	168						
58	5- 6-52	1732	4	8.5	490	2,326	1.4	*	790	280	82	87	121	65	125	13.5		
59	5- 6-52	1735	4	4.2	334	390	0.87	1,580	1,080	137	355	550	31					
60	4-28-52	1656	5	0.056	52	0.11	0.1	48,700	85									
61	4-28-52	1724	5	1.82	423	26.5	0.27	116,000	11,500	1,150								
62	4-28-52	1735	5	0.21	82	1.5	0.2	41,500	2,040									
63	4-28-52	1807	5	0.046	31	0.21	0.16	25,400	243									
64	4-28-52	1820	5	0.15	100	0.76	0.17	78,500	930									
65	4-28-52	1835	5	0.11	115	0.11	0.1	110,000										
66	5- 5-52	2038	5	0.77	250	7.4	0.20	101,000	7,030	144								
67	5- 5-52	2051	5	1.8	371	29	0.31	65,000	11,200	1,120	27							
68	5- 5-52	2117	5	0.62	208	5.2	0.2	80,800	6,400	33								
69	5- 5-52	2132	5	0.18	95	1.1	0.17	66,200	1,360									
70	5- 5-52	2150	5	0.15	156	0.15	0.1	149,000										
71	5- 5-52	2216	5	0.33	99	2.6	0.21	29,500	3,600									
72	5- 5-52	2230	5	0.55	168	4.5	0.21	51,000	6,100									
73	5- 5-52	2255	5	1.08	249	12.4	0.26	19,400	11,000	280								
74	4-29-52	1742	5	1.15	190	19.9	0.39	4,300	6,100	900	11.2							
75	4-29-52	1800	5	2.5	319	81	0.45	2,450	5,400	2,060	:70	24.2						
76	4-29-52	1813	5	2.1	381	36	0.38	16,000	13,000	1,540	19.6							
77	4-29-52	1847	5	0.17	97	0.9	0.19	71,300	1,170									
78	4-29-52	1909	5	0.66	162	11.9	0.34	10,500	6,100	480								
79	3-21-52	1645	7	1.12	141	42.7	0.47	2,380	2,050	900	110	27						
80	3-21-52	1655	7	0.11	29	1.05	0.25	2,550	1,350	4.4								
81	3-21-52	1702	7	1.9	203	87	0.59	820	1,150	1,160	400	41						
82	3-21-52	1705	7	4.44	393	475	0.69	1,470	1,780	1,290	601	159	63.6	25	5.6			
83	3-21-52	1718	7	0.24	63	2.2	0.23	8,800	2,800	12								
84	3-21-52	1727	7	1.44	190	32	0.45	7,800	2,800	1,640	34							
85	3-21-52	1745	7	1.14	160	47.5	0.4	4,600	3,950	580	94	47						
86	3-21-52	1803	7	0.51	122	4.5	0.27	4,600	6,200									
87	3-21-52	1814	7	0.061	21	0.5	0.2	8,310	666									
88	3-21-52	1837	7	0.083	54	0.4	0.17	43,000	495									
89	3-21-52	1850	7	0.12	119	0.12	0.1	114,000										
90	3-25-52	1812	7	0.9	90	68	0.59	730	170	490	134	8.2	14.3	3.1				
91	3-25-52	1813	7	2.54	212	243	0.82	640	129	600	240	195	58					
92	3-25-52	1815	7	3.1	239	443	0.91	490	400	215	390	92	94	27	4			
93	3-25-52	1825	7	1.74	154	142	0.79	640	420	400	215	170	11.3					
94	5- 1-52	1326	1	1.3	137	56	0.58	600	360	920	270	18						
95	5- 1-52	1341	1	3.6	303	430	0.83	2,000	1,500	385	390	265	52	26				
96	5- 1-52	1344	1	20.5	1,095	7,430	1.45	1,308	1,282	369	76	206	164	198	202	22.5	55.2	6.6
97	5- 1-52	1555	2	13	933	1,915	0.96	1,800	305	248	800	950	405	65	21.8			
98	5- 1-52	1608	2	3.4	374	127	0.54	534	804	3,470	436	39.6						
99	5- 1-52	1625	2	9.6	647	1,570	1.08	1,100	385	169	245	305	540	83				
100	5- 1-52	1638	2	2.8	321	97	0.53	1,010	2,950	2,350	440	10.8						
101	5- 1-52	1500	3	3	450	70	0.41	—	8,300	3,860	30.5							
102	5- 1-52	1505	3	1.1	157	25	0.43	270	2,820	1,350	12.5							
103	5- 1-52	1513	3	0.77	147	12	0.35	800	5,800	485								
104	5- 4-52	1707	3	0.75	187	7.1	0.26	20,000	8,600	56								
105	5- 4-52	1735	3	0.46	132	3.9	0.23	28,500	5,400									
106	5- 4-52	1810	1	0.95	152	19	0.39	1,358	6,320	855	21.1							
107	5- 4-52	1815	1	1.2	168	28	0.43	2,000	3,200	1,300	43							
108	5- 4-52	1820	1	1.3	214	27	0.37	4,750	6,900	971	21.1	8.4						
109	5- 4-52	1827	1	0.23	63	2.2	0.23	11,400	2,610	18.8								
110	5- 4-52	1833	1	3.6	331	277	0.73	620	1,700	850	620	225	39.5					
111	5- 4-52	1844	1	2.4	262	106	0.57	370	1,800	1,770	390	40.5	5.8					
112	5- 4-52	1846	1	24.8	1,555	5,180	1.15	—	290	196	251	710	880	* 410	85	11.2		
113	5- 4-52	1850	1	8.5	622	1,065	0.91	—	111	520	690	670	194	54				

* Splashing of large drops by wind prevented accurate determination of the concentration of these drops. It is believed that the count would have been <300 m^{-3}.

3. Measurements of drop-size distribution

Prior to the field experiments, provision was made to obtain drop-size measurements both with nylon screens and with chemically treated filter papers. In view of the difficulties encountered with the screens at high humidities, plus the fact that a low power microscope is essential for accurate determination of the drop size, the filter-paper method was adopted. An objection to using filter papers is that the papers are sensitive to changes in relative humidity (Niederdorfer, 1932). The writer found that this was especially true at relative humidities above 70 per cent. Inasmuch as the measurements of drop sizes carried out in this study were usually made at some point within the cloud, it became necessary to store the filter papers in such a manner as to keep the relative humidity below 70 per cent. This was accomplished by storing the papers in a vertical position, 6 mm apart, in a box containing several desiccating bags. Some 40 papers could be stored in this manner.

Whatman No. 1 filter papers, dusted with methylene blue dye, were held between two brass rings. These were exposed to the rain, with the aid of a small aluminum cover and a stopwatch, for any desired period of time. The exposure times, filter-paper number, time of day, and other pertinent meteorological information were recorded with pencil on painted metal strips. Data were recorded in this manner in heavy rain and cloud without any smearing whatever.

With the aid of a calibrated scale, raindrop sizes were read, in 0.2-mm intervals, directly from the filter papers. This scale was designed from a calibration curve constructed from data obtained with water drops of known size at terminal velocity. The calculation of the space distribution of the drop sizes, $N_D(\mathrm{m}^{-3}\ 0.2\ \mathrm{mm}^{-1})$, from the filter-paper distribution involves a knowledge of the effective filter-paper area ($252\ \mathrm{cm}^2$), time of exposure, drop count in each 0.2-mm size interval, and a representative terminal velocity for the drops within each size interval. The terminal velocities used in this work were those experimentally determined by Gunn and Kinzer (1949). As these velocities were determined for water drops falling in still air, it is apparent that the presence of vertical air velocities within a rain area may give rise to errors in the distribution of drops per cubic meter as computed from the filter-paper measurements. In the case of the measurements made in the orographic rain, these errors are negligible. The orographic rain samples were obtained on slopes of only 3 deg. The upslope wind was usually very low. The vertical velocity of the air, i.e., the vertical component of the upslope wind, would have little effect on the terminal velocity of the drops and, therefore, the computed spatial distributions should be correct. This, of course, applies only for the spatial distributions near the sampling area.

A rapid rate of change of terminal velocity with drop diameter is encountered with drops less than 0.2 mm in diameter.[3] For this reason, all computations of N_D for drops less than 0.2 mm are subject to error. The mass of water represented by these drops is negligibly small, when compared to the total. Therefore, computations of liquid-water content W and radar reflectivity Z are, in most cases, little affected.

The intensity of rainfall R(mm/hr) was computed from the filter-paper drop distribution. Within each 0.2-mm interval, an average mass (milligrams) was determined. This average mass, multiplied by the drop count in that particular interval, defined its contribution to the intensity. The writer realizes that such a method of determining intensities may be

[3] Unless otherwise noted, all drop sizes in this paper will be understood to be in mm diameter.

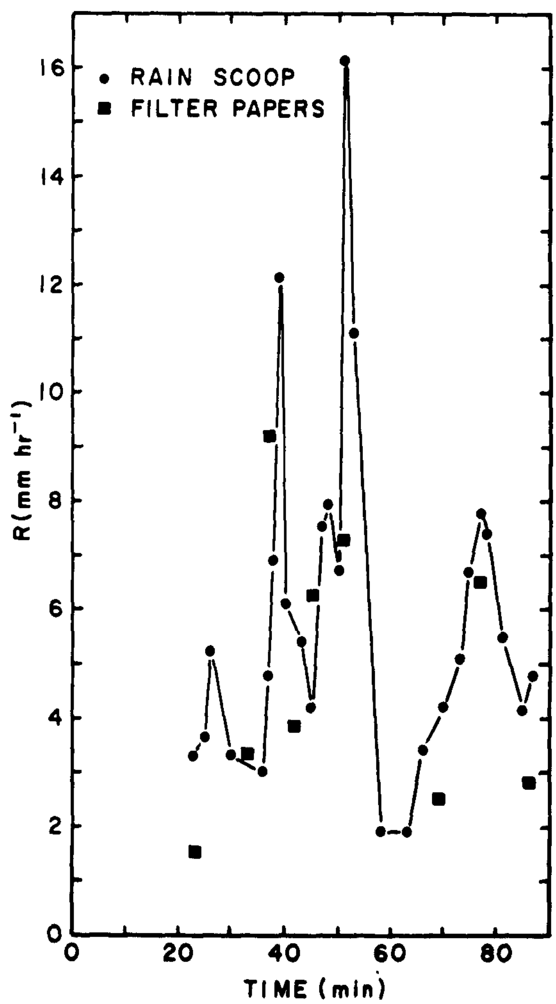

FIG. 2. Comparison of determination of rain intensity from filter papers and 80-cm diameter "rain scoop."

subject to error when the drop distributions containing large drops (>3 mm) are considered. Here the distribution of drops arriving at a horizontal surface is usually skewed, with a long tapering tail reaching into the region of large drops. In this region the distribution is often statistically inadequate and, as these large drops represent the majority of the water, incorrect intensities are computed. This is not the case, however, with the orographic rain of Hawaii. The drop-size distributions have low standard deviations, with the largest drops seldom exceeding 2 mm.

The intensities computed from filter papers have been found to agree reasonably well with those obtained with a 0.5-m^2 stainless steel funnel (see fig. 2). With the aid of a plywood cover and two flexible automobile windshield-wipers, both mounted to rotate around the inner surface of the funnel, sufficient water for intensity calculations could be collected in 10 to 200 sec. On several occasions, two such funnels were used at the same location. The results were, as expected, nearly identical. As shown in fig. 2, the average intensities as computed from funnel measurements vary considerably. The nearly instantaneous intensities computed from filter papers follow this trend probably as well as can be expected.

4. Changes in drop-size distribution in passage through sub-cloud layer

It appears that most, if not all, of the raindrop size-measurements reported in the literature were made at a considerable distance below cloud level. The changes in the spatial distribution of drops as they fall in the sub-cloud air can be considerable, depending upon the fall distance, temperature and relative humidity, relative drop sizes, and wind shear. These effects were recognized many years ago (Bentley, 1904), but received little attention as few measurements were then being made of raindrop sizes. The measurements reported in this paper, with the exception of those made in the thunderstorm and Kona storm (samples 14–30 of table 1) were obtained either at cloud base or at some point within the cloud system. This was made possible by roads which led up to elevations of many thousands of feet on both Mauna Kea and Mauna Loa.

The above-mentioned factors and their effects on the drop size distribution will be briefly discussed.

Wind shear and relative fall velocities.—If we at first consider the case of zero shear, it becomes apparent that, due to the relative fall velocities alone, large changes may occur in a spatial drop distribution between cloud and ground level. For example, consider a distribution at cloud level to contain drops ranging in size from 0.2 to 4 mm. With a cloud to ground distance of 2000 m, the 0.2-mm drops would arrive at the ground some 40 min after the 4-mm drops, with the intermediate drops arriving at successively earlier times. At ground level the distribution would be transient, not reaching the steady state until 40 min after the arrival of the largest drops. At the onset of natural rains, it is often observed that large drops precede the smaller ones by several minutes but seldom by times exceeding 10 min. This would suggest that either the drops originated at different times or positions within the cloud, or that small drops evolved as a result of continual growth and breakup of the larger drops.

If we now consider the usual case, in which horizontal winds increase with altitude, the problem becomes quite complex. It is apparent that, to have drops of several sizes arriving simultaneously at a given point on the ground, it is necessary that the large and small drops originate at different levels within the cloud or else originate at the same level with the smallest drops forming first. Both of these explanations have been considered, with the former tentatively accepted, as one explanation of observed drop distribution at the beginning stages of a shower (Atlas and Plank, 1952). However, regardless of which explanation is used, it requires that the large and small drops constituting the ground sample have their origin at different locations within the cloud.

Evaporation of raindrops.—Recent experimental work (Kinzer and Gunn, 1951) on the evaporation of falling water drops has resulted in a table of evaporation rates, at several relative humidities, for drops of various diameters. The writer has expressed this table in functional form and combined it with an expression relating terminal velocity to drop diameter. The resulting differential equation was integrated, to obtain an equation relating drop size and distance fallen. At a relative humidity of 90 per cent and an isothermal atmosphere of 20C, it was found that small drops can completely evaporate in a fall of about 1000 m. For example, a 1.5-mm drop will evaporate to only 1.42 mm in a fall of 2000 m, while a 0.5-mm drop will evaporate completely in a little over 1000 m. It is interesting to note that these calculations agree relatively well with the more detailed theoretical calculations of Best (1952).

The calculations indicate that large changes in the drop-size distribution are to be expected among the smallest drops. The evaporation of the small drops is serious, in that it deprives us of any knowledge of their distribution. This knowledge is extremely vital to the question of the mechanism of rain formation, as these drops represent the great majority of the total drops present. The great difference in numbers of small drops in rains from freezing and non-freezing clouds is pointed out later in the paper.

Drop collision in the sub-cloud layer.—As a direct consequence of the differences in fall velocities of the

various sized drops, it is to be expected that raindrop collisions in the sub-cloud layer will tend to modify the distribution which existed at cloud base. Calculations of these effects, plus those of evaporation, have been made by Rigby and Marshall (1952). They find that the collision effect tends to increase the number for large drops while decreasing it for the small ones. Evaporation effects, on the other hand, will tend to decrease the number at all sizes. On combining both evaporation and collision effects, they found that the change in distribution for the larger drops was not as pronounced as that caused by collision effects alone. The distribution of the small drops, which was decreased by both collision and evaporation, naturally departed even more from its initial state when both effects were considered. The general conclusion arrived at by Rigby and Marshall was that the basic form of the drop-size distribution would not be seriously affected by any of the aforementioned factors. It might be added that their work was based on distributions which extended into drops of 3 mm. As a majority of the drop distributions of orographic rain from warm clouds have 50 per cent of the water contained in drops smaller than 1 mm, it is to be expected that evaporation effects would be quite pronounced. In fact, the occurrence of virga, the result of evaporation, is a most common event associated with the warm clouds of Hawaii.

5. Raindrop size-distributions from clouds extending above the freezing level

On three different occasions, drop-size samples were obtained in rains whose origins most likely were associated with ice-crystal formation.

Windward Mauna Kea.—On 27 March 1952, raindrop measurements were taken on the northeast flank of Mauna Kea at an elevation of 7500 ft. These are represented by distributions 1–13 of table 1. At 0630 the weather was as follows: winds light and downslope, temperature 6.3C, and a light drizzle falling from an overcast which was solid only near the mountain. At about 0840 both the drizzle and the wind increased in intensity. Sample 2 of table 1, as compared with sample 1, shows the change in the nature of the drop distribution.[3] The absence of any drops over 1 mm and the large numbers of drops smaller than 0.5 mm in sample 1 are typical of the distributions from non-freezing clouds (see samples 31–113). The sudden increase in maximum drop size and corresponding decrease in small drops, as indicated by sample 2, was shown by all subsequent measurements until 1412 and sample 10. The change in distribution of sample 10 was no doubt associated with a wind shift to east at 1400, plus a lowering of cloud base 100 ft or more to the sampling position. At 1700 the winds became very irregular and strong. The rain continued until about 2100. At sunrise on 28 March it was observed that all of Mauna Kea above the 10,000-ft level was covered with snow. It was then realized that the rain of the previous day had probably originated as snow.

[3] Hereafter in the paper all reference to table 1 will be in terms of the sample number only.

The pronounced change in drop distribution from sample 1 to sample 2 was accompanied by a marked change in the chloride content of the rain. Chloride determinations on five rainwater samples taken between 0730 and 0842 showed the expected trend towards an inverse relationship between rain intensity and chloride content (Woodcock, 1952). During this time, the chloride concentration dropped from 20 to 0.4 ppm. From 0842 through 1802, twenty rainwater samples were obtained. Although the samples were obtained in intensities ranging from 1.6 to 13 mm/hr, the chloride concentration was never above 0.3 ppm. Rain from the typical Hawaiian orographic cloud usually has chlorides present in amounts from 0.5–20 ppm. The small amounts found above would suggest that the larger saline droplets had been eliminated from the cloud by raining out at lower elevations.

The Kona storm.[4]—Heavy and continuous rain fell throughout the day of 19 January 1952. For a period of some 20 hr, the weather was entirely dominated by a Kona or cyclonic storm. From 1031 through 1533, samples 14–22 were obtained at the Pineapple Research Institute, Honolulu. The cloud base was estimated at 200 ft. The temperature at 1200 was 20.7C, with a wet-bulb depression of 0.4C. The winds were light, with occasional strong gusts.

The drop-size measurements covered a wide range of intensities, ranging from 1.8 to 127 mm/hr. A few minutes after sample 18 was taken, the intensity rose from 127 to 242 mm/hr. This latter measurement was made with the 0.5-m^2 sampling funnel.

The thunderstorm.—On 11 February 1952, weak trade winds were indirectly responsible for the formation of cumuli over the island of Oahu. By 1300, large cumuli were forming over the city of Honolulu. Extreme vertical depth was suggested by the intense darkening of the cloud base. The first rain fell at 1352 and continued on for about 35 min. During that time, sporadic thunder was heard and small hail pellets were reported.[5]

Eight drop-size measurements (samples 23–30) were obtained. With the exception of the first three measurements, the drop distribution was, in general, similar to that found in the Kona storm. Sample 23,

[4] Kona is the Hawaiian word for leeward. A Kona storm approaches from the leeward side of the islands, with respect to the trade winds; hence, its name.

[5] According to newspaper reports, hail was reported several miles from the sampling position.

obtained 2 min after the start of the rain, contained no drops smaller than 0.8 mm. A few minutes later, at 1355, a few drops in the 0.5-mm range had arrived. At 1356 drops as small as 0.4 mm were present, although in small numbers. From 1401 on, all samples indicated the existence of drops smaller than 0.2 mm.

The drop distributions of fig. 3 show the gradual increase of small drops with time. The two dashed lines are the distributions of Laws and Parsons (1943), as presented by Marshall and Palmer (1948), for intensities of 25 and 1 mm/hr. Note how the transient is characterized by a positive slope which decreases with time. Sample 26 ($R = 8.8$ mm/hr), the first to contain drops smaller than 0.4 mm, is the first distribution that has a pronounced negative slope.

An explanation for this behavior is beset with many difficulties, arising mainly from a lack of knowledge of the drop distribution and vertical air-velocities at cloud base. With an estimated cloud to ground distance of 1000 m, and a distribution of drops of all sizes simultaneously starting their fall from cloud base, it is evident that the slower falling smaller drops will reach the ground some time after the large ones. Approximately 3 min will elapse between the arrival of drops larger than 2.4 mm and those of 0.8 mm. It is to be noted that sample 23, taken 2 min after the beginning of the rain, contains no drops smaller than 0.8 mm. Subsequent samples, obtained three or more mintues after the initial rain, contain increasing numbers of drops smaller than 0.8 mm. Thus, the time of appearance of the 0.8-mm drops agrees with the estimated time of 3 min. On the other hand, the drops smaller than 0.2 mm should not appear until some 21 min after the initial rain. Clearly this is not the case. It is, therfore, probable that the large and small drops had their origin at different altitudes, or at the same altitude but at different times (see section 4, above). The other alternatives are: (1) smaller drops are being produced by drop disintegrations resulting from collisions and turbulence (Blanchard, 1950), or (2) downdrafts created by the large drops tend to decrease the total fall time of the small drops and, consequently, decrease the elapsed time between their respective arrivals at the earths surface. In the absence of any data on the turbulence and downdrafts associated with the thunderstorm in question, it is impossible to obtain any quantitative information.

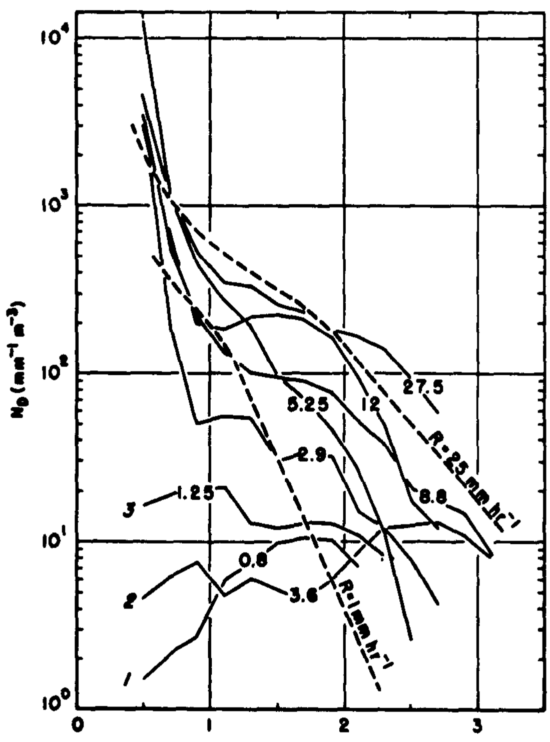

FIG. 3. Raindrop distribution (solid lines) of thunderstorm data (samples 23–30). Dashed lines represent smoothed distributions of Laws and Parsons data. Numbers in lower left-hand corner indicate the three transient distributions in a chronological order.

6. Liquid-water content as a measure of the drop distribution

It is not always convenient to compare two sets of rain measurements by comparing their drop-size distributions. It would be far more desirable to represent a drop-size distribution graphically by a single point. Of course, such a representation would tell nothing of the total drop count per cubic meter, but it could indicate whether the distribution had a large or narrow spread.

This is essentially what is measured by the liquid-water content W (mg/m^3). For example, let us consider

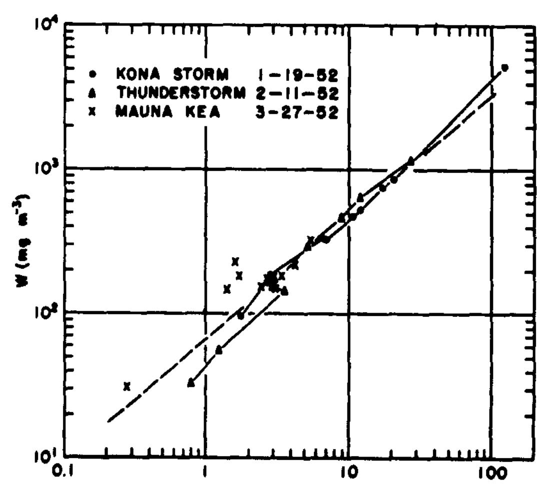

FIG. 4. Liquid-water content W as function of rain intensity R for samples 1–30. Dashed line is locus obtained by Best (1950). Lines connecting points are visual aids only.

the hypothetical distribution of 1 drop per m^3. This defines an intensity R, and a liquid-water content W. Let this drop be split into two equal-sized smaller drops. Although the liquid-water content is unchanged, the slower falling, smaller drops lower the intensity. One or more of these smaller drops will, therefore, have to be added to attain the original intensity. It is apparent that this process can be repeated indefinitely. At each sequence the intensity is held constant by adding drops, the liquid-water content rises, and the drop distribution tends toward smaller and more numerous drops.

The liquid-water contents of the drop-size distributions of the three storms represented by samples 1–30 are shown in fig. 4 as a function of the intensity R. The dashed line is the locus $W = 67\,R^{0.84}$ (Best, 1950), representing the mean value of data obtained by other investigators. With the exception of six points, the present data agree reasonably well with this locus. Note that three drop-distributions from the windward Mauna Kea rain, representing intensities less than 2 mm/hr, have liquid-water contents considerably higher than the locus would suggest. This, of course, implies a drop distribution of relatively small spread and numerous drops. Reference to samples 1, 10 and 11 shows that this is the case. In each of those samples, from 7000 to 17,000 drops per m^3 are smaller than 0.4 mm. The spread in drop distribution is about half that of the other samples.

The three anomalous thunderstorm samples indicate the opposite trend, that of a wide distribution coupled with a scarcity of small drops. These are samples 23–25, representing the transient period at the start of the storm. In each case the drop spread is greater than or equal to the other thunderstorm samples,

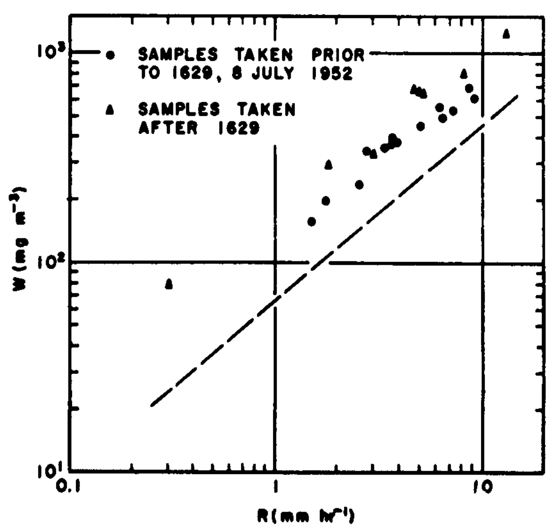

FIG. 5. W *vs.* R relationship for samples 31–52. Dashed line is locus obtained by Best (1950).

and the sum total of drops smaller than 0.8 mm, as compared to the remaining samples, is negligibly small.

7. Drop-size distributions in orographic clouds

All samples in table 1 from 31 on were obtained on the island of Hawaii, at cloud base or within the clouds. This type of sampling will eliminate or at least minimize the factors which tend to change the distribution (see section 4, above). Measurements made over a period of several months show variations in drop size for a given intensity. The liquid-water content W, as reflected by these variations, will be used to demonstrate the changes in drop distributions.

Raindrop size-distribution from clouds of windward Hawaii.—A complete overcast existed over Hilo on 8 July 1952. A light drizzle was falling at Hilo and on up through position 1 (see fig. 1). The cloud base fluctuated from ground level at position 1 to perhaps 50 ft elevation. At 1525 the temperature was 21.5C, with a wet bulb of 21.3C.

Samples 31–46 were obtained throughout a period of nearly 3 hr extending from 1423 to 1714. The trend of the drop distributions, as indicated by the liquid-water content W, is shown in fig. 5. Prior to sample 44 at 1629, the distributions fall on a common locus. All samples after 1629, with the exception of 48 and 49, have higher liquid-water contents for similar intensities. Table 1 shows this change, as might be expected, accompanied by a large increase in the drop count for drops smaller than 0.4 mm. In fact, sample 50, with its concentration of 13,700 and 4000 drops per m^4 in the 0.2-mm intervals centered on 0.3 and 0.5 mm, is one of the highest obtained by the writer. For a given R, W is 2 to 2.5 times as large as that indicated by Best's (1950) data.

Samples 53–59 were taken on 6 May 1952 at position 4. The rain-producing clouds were forming in the vicinity of the islands, as there existed an area free of

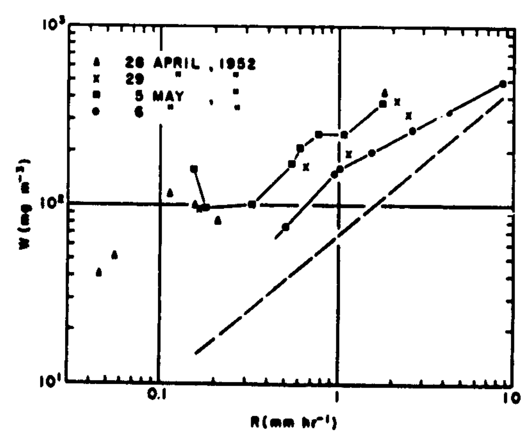

FIG. 6. W *vs.* R relationship for samples 53–78. Dashed line is locus obtained by Best (1950). Lines connecting points are visual aids only.

clouds along the east coast of Hawaii in the Hilo region. Cloud bases were approximately 100 ft above the sampling position. The dry-bulb temperature was 14.5C, and the wet bulb 14.2C. Fig. 6 shows the relative position of the W vs. R locus as compared with that of samples 60–78. Note how the points tend to approach Best's locus at the upper end.

Samples 60–65 were obtained on 28 April 1952 at position 5, at an elevation of 5500 ft. The cloud base was at 2000 ft. All but one of these samples are of intensities less than 0.2 mm/hr. With the exception of one sample, the drops are all smaller than 0.4 mm with a majority smaller than 0.2 mm. This large number of small drops can give rise to an error in the calculated intensity. This probably explains the anomalous distribution of the data.

On 5 May 1952 samples 66–73 were obtained at position 5, well within the cloud. The wind was upslope at 0.6 m/sec. The temperature was 10.8C, with a wet bulb of 10.7C. The liquid-water content for sample 70 was abnormally high. A glance at table 1 shows that all drops in sample 70 were smaller than 0.2 mm and in concentrations of 149,000 m^{-3}. This is the highest concentration of drops smaller than 0.2 mm found in the present study.

On 29 April 1952, position 5 was at or near the upper dissipating edge of the cloud. At 1640 a fine mist-like rain began to fall. The temperature was 14.4C, with a wet bulb of 13C. At 1725 the wind was steady at 1.3 m/sec. By 1740 the clouds moved in over the area, with a light drizzle which lasted throughout the time of sampling. From 1742 to 1909, samples 74–78 were collected. It is interesting to note that these distributions are similar, both in number and maximum drop size, to those obtained by Bowen (1950) from an aircraft flying through the top of a non-freezing cululus cloud.

Simultaneous with the drop-size distribution measurements at position 5, rain-intensity measurements were being made at position 4. Twenty-six measurements, from 1605 to 1905, indicated intensities ranging from 0.5 to 13.3 mm/hr. During the entire time, the cloud base was approximately at the elevation of position 4. At 1708 the dry- and wet-bulb temperatures were 16.8 and 16.7C, respectively, and at 1818 both were 15.8C.

Raindrop distributions in a dissipating orographic cloud.—In some respects, the drop-size distributions obtained on 21 March 1952 are the most interesting. For they are measurements not only made in a dissipating cloud system, but they were made at many points within the cloud system ranging from cloud base to near the cloud top.

It will be well to discuss briefly the topographical and meteorological features of the area in which this cloud forms. The region of position 7, in the lee of Mauna Kea, has little possibility of being influenced by the trade-wind flow, as is the region around positions 1–5. Leopold (1949) has shown that the 14,000-ft low-angle cones formed by Mauna Kea and Mauna Loa are sufficient to split the trade-wind flow into two components. Apparently the inversion is sufficient to prevent the flow from rising over mountains extending up through the inversion. Leopold has studied, in some detail, the formation of clouds in the lea of the 10,000-ft cone of Haleakala on the island of Maui. He found that a sea breeze was the dominant factor in the formation of the afternoon orographic clouds. In the late afternoon, this sea breeze gives way to a downslope land breeze. In many respects, we may expect a somewhat similar mechanism of cloud formation in the lee of Mauna Kea.

At 1645 on 21 March 1952, the writer was at cloud base at position 7 at an elevation of 3100 ft. The wind was nearly dead calm, and a light rain was falling. Sample 79 was taken at this point. Samples 80–83 were taken at approximately 2-mi intervals up through the cloud. Fig. 7 indicates these positions, and shows the gradual uniform rise of the slope and a schematic representation of the cloud-top positions at various times. Note the vertical structure of the cloud edge. Its 1000-ft height is based on a visual estimate.

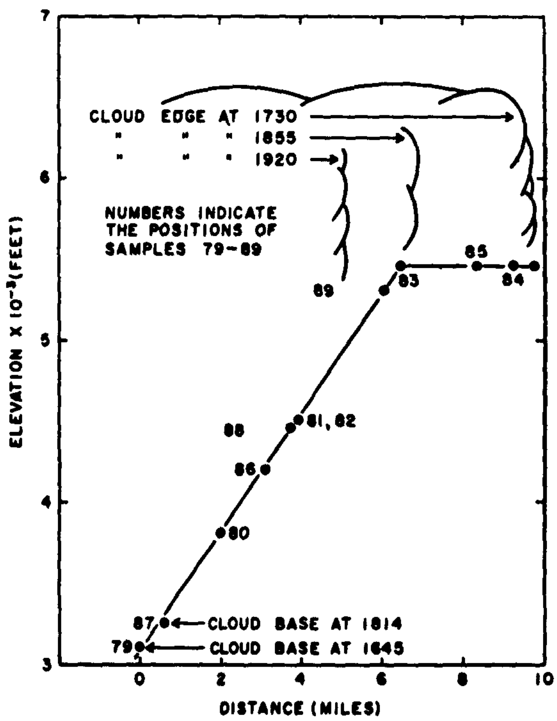

FIG. 7. Uniform slow rise of terrain (370 ft/mi) at area where samples 79–89 were obtained. Numbers alongside points indicate position of samples. Although abscissa indicates horizontal distance between points, it is to be taken as distance measured along the slope. (Actually the difference is negligible.)

Samples 85–87 were taken on the first downward traverse. During this time the cloud top was receding slowly, and the drop distribution was shifting toward the small end. This trend in the drop distribution continued during the second upward traverse, as the remaining samples, 88 and 89, were taken. From sample 86 on, a decrease was found in the number of drops in the 0.3-mm size interval and, concurrently, a steady increase in the drop count in the 0.1-mm size interval. In fact, the increase of the number of drops smaller than 0.2 mm is exponential. The equation $N = 4000\, e^{7.1\times10^{-3}t}$ can be used to express the number at t min after the time of sample 86, 1803. Within 15 min after sample 89, the cloud was void of drops of sufficient size to register on the filter paper. The apparent "drying out" of this cloud was by no means confined to these data. On other occasions the writer has been in this cloud in the early evening and has experienced the decrease in size and eventual disappearance of raindrops.

The liquid-water–intensity relationship (fig. 8) shows a fairly uniform trend with the exception of the last two samples. The large increase in W associated with these is what would be expected. Note that for the same liquid-water content of sample 89, a 17-fold increase in intensity would be required to fit Best's (1950) results.

The existence of trade-wind eddies in the lee of Mauna Kea, and high-level air flowing from east to west through the Mauna Kea-Mauna Loa saddle (Leopold, 1949), makes it very difficult to ascertain the past history of the air in this area. Woodcock's measurements have shown that significant differences in the distributions of air-borne salt particles are a function of not only wind velocities but, in some cases, of the topographical features over which the air flows.

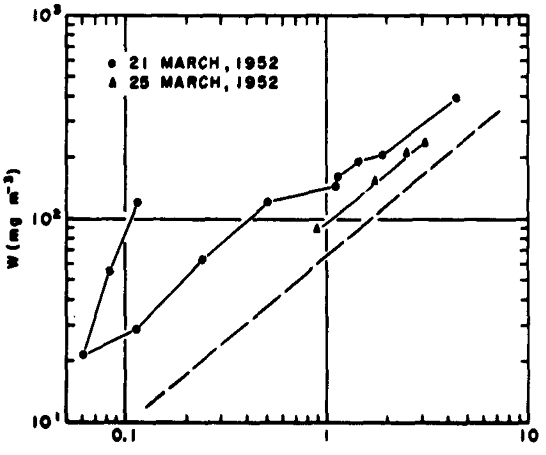

FIG. 8. W vs. R relationship for samples 79–93. Dashed line is locus obtained by Best (1950). Lines connecting points are visual aids only.

On 21 March 1952, the estimated winds to windward of the island were Beaufort force 4–5. At such speeds, the concentrations of air-borne salt particles at cloud base would be of the order of 6000–10,000 particles per m^3 between 10^3 and 10^4 $\mu\mu g$. (The equilibrium diameters of salt particles of 10^3 and 10^4 $\mu\mu g$ at a relative humidity of 99 percent are 22 and 102 μ, respectively.) And yet, on this particular day, measurements obtained from aircraft just below cloud base to leeward from Mauna Kea failed to show the existence of any salt particles heavier than 10^3 $\mu\mu g$. Ordinarily this would be typical of air only above the inversion. Whether the explanation is that this air is high-level air which has flowed down the mountain during the night, or whether it represents salt-depleted air which has passed through the saddle area from clouds on the windward side of the island, the writer cannot say. It seems apparent, however, that the presence or absence of these large salt particles should profoundly effect the rain-producing characteristics of the clouds.

Three days later, on 25 March 1952, samples 90–93 were taken at the 5500-ft level. In fig. 8 and table 1, the difference in the characteristics of the two "lee-side" distributions is obvious. A scarcity of droplets exists in the first two size-intervals. As no aircraft salt-measurements were made on this day, it is impossible to tell if the salt-particle distribution resembled that of 21 March.

Drop distributions at cloud top and base.—On 1 May 1952 a series of ten drop-distribution measurements, samples 94–103, was obtained at cloud base, an intermediate point, and near the cloud top. These are positions 1, 2 and 3 on fig. 1, with elevations of 2200, 3400, and 4000 ft, respectively. Samples 94–96, obtained at position 1, contain some of the largest drops found in orographic rain. Samples 97–100 were obtained at position 2, 6.6 mi upslope from position 1.

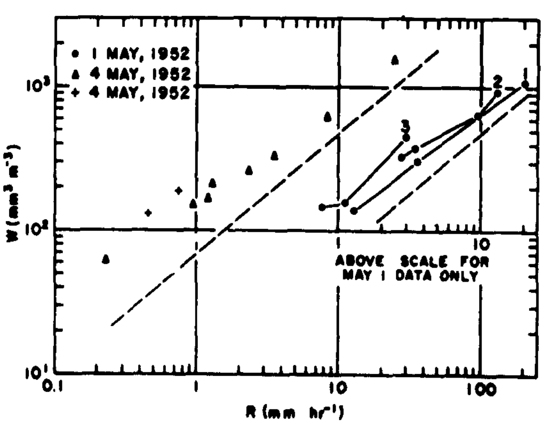

FIG. 9. W vs. R relationship for samples 94–113. Numbers alongside 1 May data indicate sampling position. 4 May data, indicated by crosses, were obtained at position 3. Dashed line is locus obtained by Best (1950).

The remaining samples were obtained at position 3, 8.8 mi upslope from position 1. At position 3, near the upper dissipating edge of the cloud, a great increase in numbers of drops between 0.2 and 0.6 mm was found. Fig. 9 shows the difference in drop distribution at the three positions, expressed in terms of the liquid-water content.

On 4 May 1952, samples 104–113 were obtained at positions 1 and 3. At 1730, at position 3, both wet- and dry-bulb readings were 13.4C. The difference in drop distribution between the two positions is illustrated in fig. 9. This difference becomes numerically clear by inspection of table 1. The large number of drops smaller than 0.4 mm is sufficient to cause a high liquid-water content.

8. Median volume diameter as a function of rain intensity

Many of the data of table 1 have been expressed in fig. 10 in terms of median volume diameter d_0. The d_0 is that diameter which divides the drop distribution into two parts, such that each represents half of the liquid-water content W. It is obtained by plotting a cumulative per-cent curve of the liquid-water content. The percentage corresponding to any drop diameter is the percentage of the total liquid-water content contained in the drops smaller than the drop in question. The drop diameter at the 50 percent ordinate is, therefore, the median drop diameter d_0.

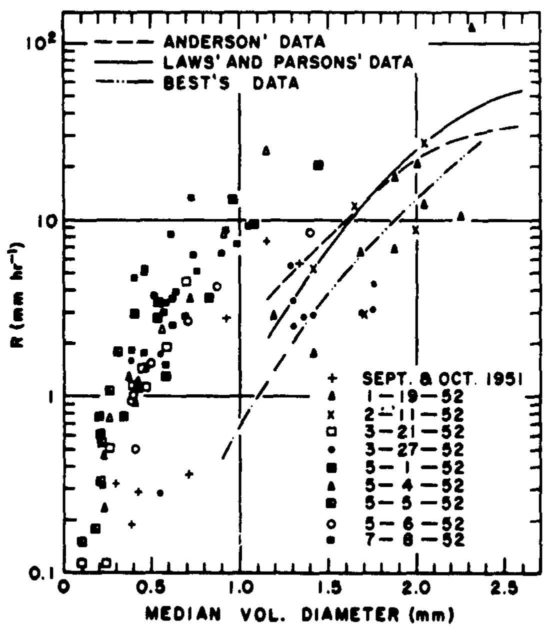

FIG. 10. Median volume diameter as related to rain intensity. Data labeled Sept.–Oct. 1951 were obtained at Woods Hole, Mass. Median volume diameters, as found in orographic rains, are considerably less than those found in non-orographic rains.

In addition to the data from the Hawaiian orographic rains, data from non-orographic rains (samples 1–30 and drop distributions obtained at Woods Hole, Massachusetts) have been included in fig. 10. With the exception of four of the samples from the Mauna Kea and Woods Hole data, all the median diameters greatly exceed those found in orographic rains of the same intensity. Note that the three Mauna Kea samples (1, 10 and 11 in table 1) which fall into the orographic grouping have drop distributions representative of orographic rains. The median diameters of the non-orographic samples alone show considerable spread at all intensities. In view of the differences already pointed out in the $W-R$ relationship, this spread is to be expected.

The solid line was drawn from the data of Laws and Parsons (1943), the dashed line from the data of Anderson (1948), and the dash-dot line from the data of Best (1950). Laws and Parsons used the flour technique for drop-size sampling (Bentley, 1904) and calculated the intensity of rainfall from the exposures, area, and drop distribution of the sample. All of their rain samples were obtained at ground level at Washington, D. C.

Anderson's results are extremely interesting, in that they were taken on the island of Hawaii in the vicinity of position 4 (fig. 1). Some 60 samples were obtained with the blotting-paper method over a period of 5 hr. The disagreement of Anderson's data with the present data, and the relatively good fit with that of Laws and Parsons, and Best, suggests that Anderson's sampling was in a particular rain not representative of the general Hawaiian rains. Of the 60 samples, only three were taken at intensities less than 8 mm/hr and none at intensities less than 2.5 mm/hr. Anderson states,[6] however, that the rain appeared to be orographic in nature and was accompanied by light winds. Nevertheless, it is possible that this rain was similar in origin to that of samples 1–13, evolving from snow from high-level supercooled clouds. From a meteorological point of view, this was quite possible. Anderson's work was carried out on 16 March 1945, the same time of the year as samples 1–13. During the winter months, and extending through March, it is not an infrequent occurrence to have rain of this nature.

The quartile deviation for orographic rain, a measure of the spread of the liquid-water content, is considerably lower than that reported by Anderson.

The present data indicate values ranging from 0.01 to 0.15 mm, as compared to Anderson's measurements of 0.1 to 0.8 mm. The writer finds, as did Anderson, that the quartile deviation is roughly proportional to the median diameter. This, of course, implies a decreasing slope of the cumulative per-cent curve

[6] Private communication.

between the first and third quartiles with increasing median diameter. According to Anderson, this is contrary to the cumulative per-cent curves of Laws and Parsons, which show a nearly constant slope between the first and third quartiles at all median diameters.

9. Radar reflectivity

The success of radar in determining the intensity of precipitation is dependent on a knowledge of the size distribution of the precipitation elements. The power received at a radar from a rain target is proportional to the radar reflectivity $Z = ND^6 \delta D$, where N is the number of drops per cubic meter of diameter D in the size interval δD. It is apparent that the sixth-power-of-the-diameter factor allows the relatively few large drops greatly to influence the radar reflectivity.

Wexler (1948) and Marshall and Palmer (1948) have computed, from their own data and that of other investigators, the relationship between Z and rain intensity R. Recently, similar relationships have been found to hold for various spectrums of cloud droplets (Atlas and Boucher, 1952). Marshall and Gunn (1952) report that the Z *versus* R relation has been found to be interchangeable for rain and snow. That is, for equal rates of precipitation, whether rain or snow, they obtain the same values of Z. Twomey (1953) has presented the results of Z calculations made in Australia. He points out that, for a given intensity, the drop-size distribution may vary considerably. This, of course, implies a corresponding variation in Z. Twomey has presented a list of $Z-R$ equations obtained by many investigators at widely separated localities. Considerable disagreement exists in these equations. They range from $Z = 23.5\ R^{2.028}$ to $Z = 1600\ R^{1.4}$. The Australian results alone indicate that the rain intensity, as deduced by radar, may be in error by a factor as great as four.

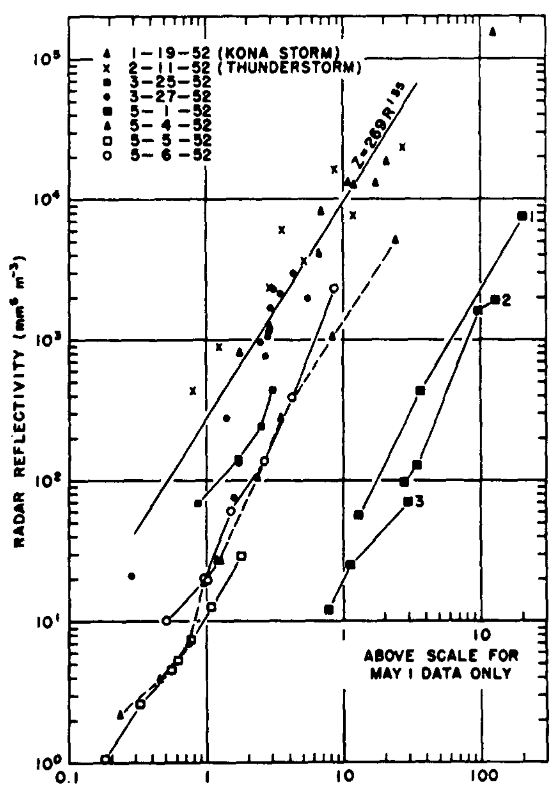

FIG. 11. Radar reflectivity as function of rain intensity. Locus $Z = 269\ R^{1.55}$ is not that of present data, but one obtained by staff of Mt. Washington Observatory in least-squares analysis of Z *vs.* R relation of continuous and shower-type rain at Cambridge, Mass. Numbers connected with 1 May data represent sampling positions. Lines connecting points are visual aids only.

These variations in the $Z-R$ equations are not surprising. They undoubtedly represent rains whose origins lie in snow-producing clouds, non-freezing cumuliform clouds, and orographic-type clouds. Further variations are probably introduced by the evaporation and collision of drops in the sub-cloud region (see section 4, above). Fig. 11 shows how the Hawaii data alone vary in Z for a given R. The $Z-R$ relationship for samples 1–30, the non-orographic rains, most nearly corresponds with that of other workers. The regression line shown was not drawn on the basis of these data. It represents the least-squares regression of 63 rain samples, both from continuous and shower-type rain, taken at Cambridge, Massachusetts (Mt. Washington Observatory, 1951b). The regression for the three types of rain represented by samples 1–30, excluding samples 1 and 10, is $Z = 290\ R^{1.41}$. Samples 1 and 10, as seen from table 1 and discussed in section 5, above, are representative of orographic rain and would, therefore, show relatively low values of Z.

The Hawaiian orographic rains of low intensity (less than 2 mm/hr), as compared with the non-orographic rains, may give lower values of Z by as much as a factor of 30.[7] At intensities greater than 10 mm/hr a factor of from four to ten is found. Inasmuch as day-by-day variations exist, it is not felt necessary to present any least-square fits. However, it can be easily seen from inspection of fig. 11 that the coefficient in the $Z-R$ equation will be from 10 to 100, considerably lower than those found elsewhere. The data of 1 May 1952, obtained at three different positions within the cloud, illustrate the small but noticeable difference in Z in various parts of the cloud.

If the type of rain, *i.e.*, thunderstorm, frontal, orographic, is not known, a large error may be made

[7] A factor of 15 is probably the usual case. The factor of 30 was based on the 5 May 1952 data, obtained at position 5. Only in the case of a subsiding cloud (samples 79–89) would one expect to find such a drop distribution at cloud base.

in determining R by radar. In agreement with the Australian findings, the error indicated here may be as large as a factor of four.

10. Generalized nature of the drop distribution

The variations in the drop distributions.—The discussions above, and figs. 5–9, have dealt with the changing drop-distributions that exist from day to day and from one position to the next. The writer feels that an adequate explanation for these apparent anomalies is to be found only by considering the past history of the cloud. For example, one would want to know just how long a cloud system has been raining prior to the time that the raindrop samples are obtained. Some of the present samples were obtained when it was observed that the rain area extended many miles to windward. On the other hand, other samples were obtained at or near the beginning of the rain area. If large airborne salt-particles are the nuclei upon which the raindrops form (Woodcock, 1952), it is to be expected that these particles will rain out as the cloud system advances and, therefore, provide a variation in the drop distribution.

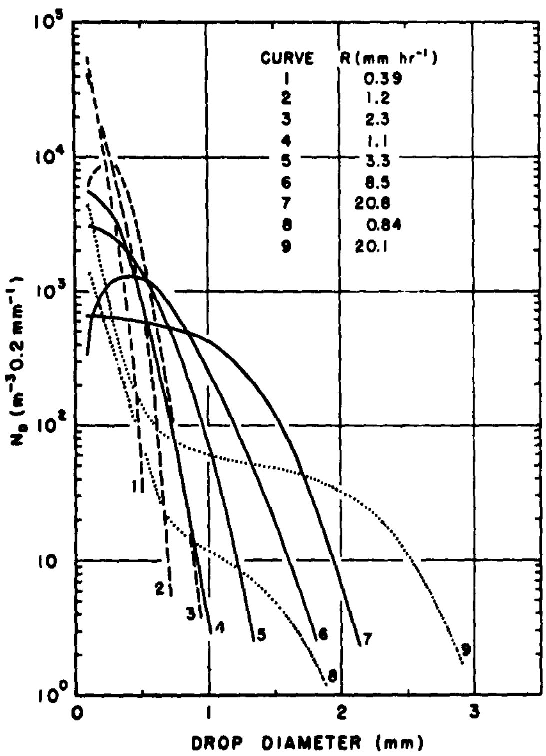

FIG. 12. Raindrop distributions, as averaged from data of table 1. Curves 1–3 are for measurements made at or near dissipating edge of non-freezing orographic clouds (positions 3 and 5), while curves 4–7 represent data taken at cloud base (positions 1 and 4). Curves 8–9 are for non-orographic rain distributions (samples 1–30). It should be noted that these curves were drawn for best fit to the averaged data. In all cases, with exception of curve 7, little or no "smoothing" of curves was required. In case of curve 7, however, data showed considerable scatter about "smoothed" line.

Fig. 12 represents, then, an average drop-size distribution around which systematic fluctuations can and do occur. The three dashed curves represent the averaged distributions of positions 3 and 5, while the solid curves are for distributions obtained at positions 1 and 4. It is immediately apparent that the distributions well up in the cloud (positions 3 and 5) are markedly different, for the same rain intensity R, than those at cloud base (positions 1 and 4). For example, curves 2 and 4, representing nearly identical intensities of 1.2 and 1.1 mm/hr, respectively, show large differences at each end of the drop spectrum. The number of drops smaller than 0.2 mm is nearly a factor of ten greater within the cloud than at cloud base. On the other hand, the number of drops per cubic meter between 0.6 and 0.8 mm is a factor of ten less within the cloud than at cloud base. A somewhat similar picture is presented by curves 3 and 5.

A second feature of these curves is that, in general, the number of drops per cubic meter at the small end of the spectrum is an inverse function of the intensity. It is believed that the scarcity of small drops at high intensities is due to accretion with the large drops. It will be noted that the numbers of large drops, as well as the maximum size, increase with intensity.

The two dotted curves are for some of the data from the non-orographic rains (samples 1–30). In all respects, they are markedly different from the other curves. The inverse relationship of intensity *vs.* raindrop concentration at the small end of the spectrum is not obtained, and the curves exhibit a more uniform distribution of drops beyond 0.7 mm. Without further data on this type of rain, one can only speculate as to why these curves differ from those of the orographic rain. As it is most probable that these curves represent rain evolving from snow falling through the freezing level, it is likely that the size distribution of the snowflakes and the manner in which they melt determine the basic shape of the raindrop distribution curve.

If one closely considers the nature of the drop distribution curves for the orographic rain, it becomes apparent that Langmuir's (1948) "chain reaction" process does not take place. This hypothesis postulates the existence of updrafts and cloud thicknesses which must exceed a critical value. Raindrops are presumed to grow to a point where turbulence or drop collision causes breakup into two or more smaller drops which, in turn, repeat the same process. This cannot occur here, for, in the first place, vertical velocities and cloud thicknesses of the order necessary for chain reaction are not observed in Hawaiian non-freezing orographic clouds. Secondly, and perhaps more important, a necessary condition for chain reaction is

the presence of drops of some 5 mm.[8] The absence of drops larger than 2 mm is quite apparent in the present data. Any chain-reaction process would certainly provide a spectrum of drops between 2 and 5 mm. It is much more likely that the raindrops are the product of a simple accretional growth process. The evidence in support of this process is being prepared for publication at the present time.

Regressions equations and determination of the median volume diameter.—The important meteorological parameters, liquid-water content W, radar reflectivity Z, and the median volume diameter d_0, have been determined for the distributions of fig. 12, and have been plotted against the intensity R. The regression equations relating these data are given in table 2. Note that the Z regression equation for the non-orographic rain is in good agreement with that of fig. 11, while the equation for W is similar to that given by Best (1950), $W = 67\ R^{0.87}$. The regression equations for the orographic rain are, as must be expected, considerably different than those for the non-orographic rain. For example, the coefficients for the three Z equations vary by a factor of 17.

[8] It is at about this size that raindrops are thought to become unstable. The writer (Blanchard, 1950) has observed that water-drops up to 8 mm are stable in non-turbulent air. It is only when an 8-mm drop encounters a high degree of turbulence that breakup will occur. This turbulence may be part of the air structure itself, or it may be set up by the aerodynamic interaction of two drops about to collide. The droplets resulting from the breakup, in either case, will have a variety of sizes. Drops of only 3 or 4 mm, on the other hand, are stable even when falling in turbulent air or when colliding amongst themselves.

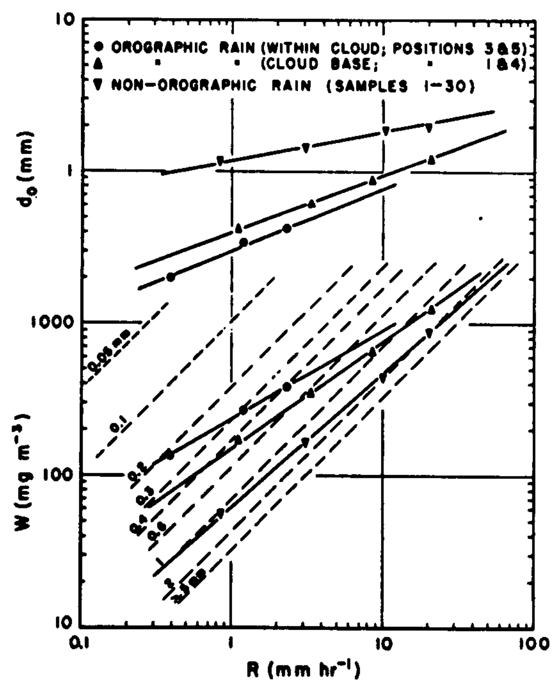

Fig. 13. Average distributions of fig. 12 represented as regression lines for parameters d_0 (upper set of lines) and W (lower set of lines) as function of R. Dashed lines on W-R graph constitute a family of lines of uniform drop size. Subject to standard deviations given in text, these lines may be considered as lines representing d_0.

Table 2. Regression equations for raindrop-distribution parameters W, d_0 and Z, for both orographic and non-orographic rains. (R in mm/hr.)

Type of rain	W (mg/m³)	d_0 (mm)	Z (mm⁶/m³)
Orographic rain within the cloud (positions 3 and 5)	$235\ R^{0.57}$	$0.302\ R^{0.4}$	$16.6\ R^{1.88}$
Orographic rain at cloud base (positions 1 and 4)	$150\ R^{0.70}$	$0.397\ R^{0.27}$	$31\ R^{1.71}$
Non-orographic rain (Samples 1–30)	$61\ R^{0.90}$	$1.175\ R^{0.19}$	$290\ R^{1.41}$

The d_0 and W data from the averaged distributions are shown in fig. 13 as functions of R. The dashed lines, with a slope of one, are for a collection of drops of uniform size. These lines illustrate that, for any given intensity, the liquid-water content will increase as the drop size decreases (this was discussed in section 6, above). The plotting of these lines on the same graph as the regression lines enables one to make a reasonably accurate estimate as to the median volume diameter d_0 of the drop distribution. As an example, consider the regression line for the orographic rain within the cloud. At an intensity of 1 mm/hr, the liquid-water content is 235 mg/m³. The dashed line for drops of 0.3 mm crosses the regression line at this point. This value of 0.3 mm is identical with the value of d_0 found with the aid of the upper group of lines of fig. 13. At an intensity of 2.1 mm/hr, the regression line crosses the line for drops of 0.4 mm. Again this is seen to coincide with the actual d_0. It is interesting to note that the dashed lines reach a limiting position at a drop size of about 5 mm. At this size, the terminal velocity has effectively reached a maximum.[9] A further increase of drop size, say to 10 mm, will give rise to an eight-fold increase in W and, as the terminal velocity remains unchanged, a similar increase in R. Thus, the new position on the $W-R$ graph will still be on the line for 5-mm drops. If any drop distribution has values of W and R which locate it to the right of the line for 5-mm drops, a mistake in the calculations of W or R is implied, or else terminal velocities exceeding those with respect to a stationary reference must be inferred. The latter case presumably could be found in downdrafts or at high altitudes.

The median volume diameters were individually determined for each of the drop samples of table 1. These were compared with the value estimated by the method explained above. The standard deviation of d_0 for the non-orographic rain (samples (1–30) was 15.6 per cent and, for the orographic rain, only 11 per cent. It is interesting that, for the non-orographic

[9] In reality, the terminal velocity continues to increase with drop size beyond 5 mm. This increase, less than 1 per cent, can be neglected in the present study.

samples, the estimated d_0 was less than the actual d_0 in about 75 per cent of the cases, while for the orographic rain the estimated d_0 exceeded the actual d_0 in 70 per cent of the cases. Regardless of this, the relatively small standard deviation allows one to use the dashed lines of fig. 13 as lines of constant d_0.[10]

The validity of using the lines of constant drop size as lines of constant d_0 was confirmed even for cloud-droplet distributions. The parameters R, W, Z and d_0 were obtained for the nine cloud-droplet distributions published by Squires and Gillespie (1952). With the exception of the first three distributions, which extended well into the drizzle range, the standard deviation of the estimated d_0 was only 4.2 per cent. These distributions gave intensities from 0.1 to 0.4 mm/hr and liquid-water contents in the region of 1000 mg/m^3.

The function $Z = f(d_0, W)$, identical for cloud and raindrop distributions.—In a paper on the reflection and transmission characteristics of microwaves in clouds, Bartnoff and Atlas (1951) presented the equation $Z = 6\pi^{-1} G(n)\, d_0{}^3 W$, where $G(n)$ was a factor depending on the spread of the cloud-drop distribution. In a later paper (Atlas and Boucher, 1952) over 100 cloud samples were analyzed, with the result that $G(n)$ could be taken as a constant of 1.35 with a standard deviation from regression of 35 per cent.

The value $d_0{}^3W$ was computed for each distribution of the present study and plotted against Z.[11] The regression line was calculated to be $Z = 6\pi^{-1}\,(1.37)\, d_0{}^3 W$, with a standard deviation from regression of 30 per cent for the coefficient. The exponent of $d_0{}^3W$ was 1.007, near enough to unity to insure the reliability of the exponents 3 and 1 on d_0 and W, respectively. The cloud-drop distributions (Squires and Gillespie, 1952) gave a regression line of $Z = 6\pi^{-1}\,(1.32)\, d_0{}^3W$, with a standard deviation from regression of 13 per cent for the coefficient. The good agreement of these three regression equations indicates uniqueness for cloud- and raindrop-distributions alike.

11. Summary and conclusion

1. A given drop-size distribution can be modified by wind shear, relative fall among the drops, evaporation, and drop coalescence in the fall from cloud to ground. Although some of these factors are at work within the cloud itself, it is certain that drop-size sampling at the cloud base will minimize the errors contributed by these factors. The evaporation will be most important, especially in the case of the semi-tropical orographic rains discussed in the present study. In these rains, the many thousands of drops per cubic meter smaller than 0.5 mm that are normally present may evaporate completely in a sub-cloud fall of 1000 m. This evaporation was eliminated in the present work by obtaining the orographic drop distributions on the sides of the volcanoes of the island of Hawaii at cloud base or within the cloud itself.

2. Drop distributions have been obtained in rain which presumably began as snow in freezing clouds. The differences in drop distribution, liquid-water content, median volume diameter, and radar reflectivity from that of orographic rains are apparent from table 1 and figs. 4–11. The liquid-water content W has been used as a measure of the drop distribution. A wide distribution with relatively few drops, both large and small, will give a lower value of W, for the same intensity, as will a narrow distribution composed of many small drops. W-R relationships for non-orographic rains have been found to agree reasonably well with that of Best (1950).

3. The distribution of raindrops in semi-tropical, non-freezing orographic clouds is decidedly different from the drop distributions presented in the literature. In general, the number of drops per cubic meter at the small end of the raindrop spectrum is an inverse function of the rain intensity. The number of large drops is a direct function of the intensity. The maximum drop size seldom exceeds 2 mm and concentrations of drops smaller than 0.5 mm often exceeds 25,000 per m^3. These distributions of drops are not the result of chain reaction. It seems probable that these drops evolve first by condensation on large air-borne salt particles, and then by accretional processes with the numerous cloud droplets. The evidence for this hypothesis will be presented in the near future. The raindrop distribution near the top of orographic clouds is concentrated at the small end of the spectrum. The appearance of drops larger than 0.6 mm is exceptional.

4. The median volume diameter, the drop diameter at which the total volume of water per cubic meter is divided equally, has been presented as a function of the intensity of rainfall. For a given intensity in an orographic rain, the median volume diameter is about half that found in thunderstorm and frontal type rains.

5. The radar reflectivity Z in an orographic rain is a factor of 10–20 less than that found in thunderstorm-type rains. Variations of Z have been found in orographic rain from day to day.

6. Regression equations for the parameters d_0, W and Z as a function of R illustrate the basic differences between orographic rain from non-freezing clouds and the type of rain which develops by the Bergeron-Findeisen process. The regression $Z = f(d_0, W)$ is identical for all the drop distributions of the present study and for the cloud distributions of Squires and Gillespie (1952). The median volume diameter d_0 can be found with reasonable accuracy from a W-R plot of a family of lines of constant drop size. This indicates that, in general, a drop distribution can be represented by a uniform collection of drops with a drop size equal to d_0. The parameters W, Z and R will be unchanged in either case.

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[10] A somewhat similar method for determining d_0 for cloud-droplet distributions was used by Atlas and Boucher (1952), by using a W-Z graph.

[11] The Z values for the data ranged from 10^{-1} to 10^4. Some ten or more points with $Z < 3$ were not used in computing the regression line. As most of these points represented drop samples with a spread of only 0.2 to 0.4 mm, it was felt that the small errors in determining the drop diameters would be magnified in the calculation of Z.

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